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Citation for published version:
Kaplan, MR, Hein, AS, Hubbard, A & Lax, SM 2009, 'Can glacial erosion limit the extent of glaciation?', *Geomorphology*, vol. 103, no. 2, pp. 172-179. https://doi.org/10.1016/j.geomorph.2008.04.020

Digital Object Identifier (DOI):
10.1016/j.geomorph.2008.04.020

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Abstract

In southern South America, the maximum areal extent of ice during the Quaternary Period, the Greatest Patagonian Glaciation (GPG, Mercer, 1983), occurred at 1.1 Ma and subsequent glaciations were overall less extensive. The GPG preceded global minimum temperatures and maximal volume of ice, which occurred in the last ~ 800 kyr, as recorded in the marine $\delta^{18}O$ record. Significant modification of the drainage morphology of the southern Andes from a non-glaciated to glaciated landscape occurred throughout the Quaternary Period. We infer a non-climatic relationship between glacial modification of the mountains and decreasing the extent of ice and we discuss processes of landscape development that could have caused the trend. Specifically, these include modification of valleys, such as development from a V- to a U-shape, and lowering of mass-accumulation areas. Such changes would strongly affect glacial dynamics, the mass balance profile and mass-flux during succeeding glaciations, especially for low-gradient outlet glaciers occupying low areas. Other areas around Earth (at least where ice has been warm-based) also may exhibit a non-random trend of decreasing the extent of ice with time, ultimately because of glacial erosion in the Quaternary Period.

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1. Introduction

In southern South America, chronologic data indicate that the greatest areal coverage of ice (Greatest Patagonian Glaciation: GPG) occurred prior to 1 Ma and, overall, succeeding glaciations have been less extensive (Figs. 1, 2). The trend exists for > 1500 km along the southern Andes, it is not particular to any individual valley, and the successions contain several moraines of different ages (Rabassa and Clapperton, 1990; Mercer, 1983; Singer et al., 2004a). Gibbons et al. (1984) showed that the expected preservation is only 2-3 moraines for a random succession of advances, given that subsequent glacial advances remove previously deposited moraines. More than three moraines (from different glaciations) and represented in various areas studied along the Andes (Clapperton, 1993; Meglioli, 1992; Rabassa et al., 2000; Singer et al., 2004a; Kaplan et al., 2005) point to a non-random explanation for the overall trend. The pattern is also observed along most of Patagonia despite differences in tectonic setting and modern climate (Prohaska, 1976; Ramos, 2005). Determining whether this trend in decreasing ice extent occurs because of climate and/or other effects, such as topographic changes or tectonic processes, is relevant to understand the general history of the Andes.

The decreasing trend of coverage of Patagonian ice does not mirror Southern Ocean temperature changes over the last 1 Myr (e.g., Hodell et al., 2002; Carter and Gammon, 2004). Although much work remains to be done in documenting the paleotemperature of the Southern Ocean over this time, no known observed trend (in records that resolve glacial-interglacial cycles) is consistent with the decreasing extent of Andean ice over the last 1 Myr. In addition, decreasing coverage of Patagonian ice differs from a key feature of the benthic $\delta^{18}O$ record, which suggests that volumetrically, the ice on the planet (i.e., in the Northern Hemisphere) reached a maximum after 0.8 Ma or in the Middle to Late Quaternary (Fig. 3), > 200 kyr after the

GPG. The most extensive ice in southern South America occurred before the 100 kyr Milankovitch periodicity and was the dominant pacing of the largest volume (and areal?) changes of the Northern Hemisphere ice sheet. Perhaps the greatest volume of Andean ice occurred over the last ~ 800 kyr, coincident with the global volume of ice, but differed from the maximum areal extent during the GPG. Our focus here is on changes in areal extent of glacial ice.

If climate change is not an obvious cause of the GPG and subsequent less extensive Andean ice, then other explanations need to be invoked. In the Quaternary, glaciers have substantially modified the landscape (Sugden and John, 1976), especially in temperate middle latitude areas such as Patagonia where basal conditions were warm-based. Here we argue for a connection between the phenomena of glacial erosion and the pattern of the extent of ice over time in southern South America, which builds upon prior discussions of the topic (e.g., Rabassa and Clapperton, 1990; Singer et al., 2004a). We focus on the records of glacial landforms in the Andes and possible changes in paleoglaciology, in comparison to other papers on the subject of glaciations and landscape development (e.g., Harbor, 1992; Small and Anderson, 1998; Whipple et al., 1999; MacGregor et al., 2000; Montgomery et al., 2001; Tomkins and Braun, 2002; Thomson, 2002). We propose that a 'self-defeating mechanism' (MacGregor et al., 2000, p. 1033), whereby glacial erosion causes less extensive coverage of ice over time, and plays a key role in the development of the Andes, and perhaps elsewhere.

2. Decreasing ice extent in Patagonia

We focus on southern South America to observe long-term changes in the extent of ice because (in places) it contains one of the best-dated, longest glacial moraine (i.e., frontal position) records on Earth, ~ 1 Myr (Mercer, 1983; Rabassa and Clapperton, 1990; Clapperton, 1993). From 40ºS to 55ºS latitude, moraines and till sheets clearly show that glacial expansions have been less extensive in general since the GPG, at 1.1 Ma (Caldenius, 1932, Rabassa and Clapperton, 1990; Mercer, 1983; Meglioli, 1992; Singer et al., 2004a; Coronato et al., 2004a; Rabassa et al., 2005). Lava flows interbedded with glacial deposits allow a multi-chronologic approach to dating the record; radioisotopic and cosmogenic nuclide data provide quantitative ages on at least a broad chronologic framework for the last 1 Myr (Singer et al., 2004a, b; Kaplan et al., 2005).

The GPG and successive glaciations are best dated in two areas of southern Patagonia. At Lago Buenos Aires, Argentina (Figs. 1, 2), glacial events over the last 1 Myr are defined in age with $^{40}Ar/^{39}Ar$, K-Ar, ^{10}Be, ^{26}Al, ^{3}He, and radiocarbon data (Singer et al., 2004a; Kaplan et al., 2005; Douglass et al., 2006). The oldest glacial deposit, underlying the Arroyo Telken Flow (Fig. 2), is the local representative of the GPG and is dated to >1.016 Ma (Singer et al., 2004a). The $^{40}Ar/^{39}Ar$ and K-Ar chronology (Singer et al., 2004a) of three lava flows (Fig. 2) indicates that at least six Telken moraines (informal names) were deposited between 1016 and 760 ka, five Deseado/Moreno moraines between 760 and 109 ka, and six Fenix moraines after 109 ka. Cosmogenic nuclide data further define the ages of the five Moreno/Deseado moraines between 760-109 ka; at least two of the moraines date to ~150-140 ka, and the other older three are likely >300 ka (Kaplan et al., 2005; Douglass, 2005). Kaplan et al. (2004) and Douglass et al. (2006) define further the youngest innermost set of moraines adjacent to the lake to between ~ 23 and 14 ka or marine isotope stage 2. At Lago Buenos Aires, the GPG deposit is ~ 50 km farther east than the last glaciation stage 2 moraines (Fig. 2). The GPG and successive glaciations are also well-dated, at least for the purposes of this study, in southernmost Patagonia (e.g., for the Bella Vista lobe), where $^{40}Ar/^{39}Ar$ ages of 1.168 and 1.07 Ma locally provide maximum and minimum ages of the GPG (Meglioli, 1992; Singer et al., 2004a). Subsequent glaciations in southernmost Patagonia/Tierra del Fuego are further defined with $^{40}Ar/^{39}Ar$, U-Series, and cosmogenic nuclide data (Meglioli, 1992; Rabassa et al., 2000; Kaplan et al., 2007).

For other areas of Patagonia, including Tierra del Fuego, Rabassa et al. (2000) and Coronato et al. (2004b) summarized the history of glaciations, including the geochronologic data. On the west side of the southern Andes, south of Isla Grande de Chiloé, the limit of the GPG and subsequent glaciations may have been always the continental shelf break (Clapperton, 1993; Hulton et al., 2002; Hubbard et al., 2005). Although numerical data are lacking for parts of Patagonia, mapping over the last 100 years has traced ice margin deposits along Patagonia between the key more recent chronologic efforts discussed above and on the west side of the Andes north of Chiloé, which all support the existence of at least the broad temporal pattern (Nordenskjöld, 1899; Caldenius, 1932; Clapperton, 1993; Coronato et al., 2004a,b).

The areal extent of GPG ice in southern Patagonia is estimated to be ~558,000-542,000 km^2, compared to ~ 422,000 km^2 during the Last Glacial Maximum (LGM Fig. 1). These estimates were arrived at by digitizing the mapped extent according the references cited above and using a geographic information system (Fig 1B). During the GPG, ice appears to have formed a continuous ice sheet along the Andes south of ~ 50°S, where its eastern outlet lobes overran a landscape of piedmont gravels, which was less dissected than at present (Rabassa and Clapperton, 1990; Singer et al., 2004b). Although some moraines in any particular valley may reflect a random succession of advances between the GPG and the last glaciation, recent chronologic campaigns, building upon detailed mapping efforts, allow the following confident assumptions: 1) the greatest extent of ice in the Quaternary was at the GPG; 2) the most recent glaciation was far less extensive, and 3) more than two glaciations are preserved between 1 Ma and ~ 25 ka.

3. Glacial modification of the landscape

Glacial coverage causes marked changes to a landscape. Prior to glaciations a landscape contains a fluvial drainage network (e.g., V-shaped valleys). For glaciations to occur a region must have sufficient area above the snowline for ice to accumulate. Over time landscapes of alpine glaciation, areal scouring, and selective linear erosion can develop, depending on the topographic and climatic setting. Alpine glaciated landscapes are obviously limited to mountainous areas such as in the Andes. We refer to definitions in Sugden and John (1976): "landscapes of areal scouring bear signs of pervasive glacial erosion; selective linear erosion describes situations where ice erosion has been concentrated in low areas leaving intervening uplands unmodified; and alpine glaciated landscapes exhibit a dendritic network of troughs with steep, often precipitous slopes" (pgs. 194-197). An obvious example of a geomorphic change is the widening and deepening of valleys and fiords and conversion to troughs with an increased cross-sectional area to perimeter (U-shaped valleys), which more efficiently discharge ice. Many excellent examples in the literature discuss and model the above processes (e.g., Harbor, 1992; MacGregor et al., 2000; Montgomery, 2002). Focusing on the southern Andes, we discuss potentially important effects, which are not independent, that landscape modification could have on glacial activity.

First, topographic changes will have important implications for glacial dynamics, especially if a glacial system preferentially erodes low-lying areas (e.g., Jamieson et al., in press). Development of a landscape of selective linear erosion will significantly alter the style of glacial activity. Modification of low areas that efficiently discharge ice will yield more rapid movement and a tendency for fast-flowing outlet glaciers (bounded by bedrock) and ice streams (bounded by ice). As large sectors of the ice sheet and outlet glaciers evolve towards fast-flow with lower surface slopes, the response time increases and they could react more quickly to changes in mass balance. Selective linear erosion can lead to a landscape that facilitates contrasting basal regimes that include cold-based (on uplands) and warm-based (e.g., valleys and fiords) bed conditions, especially in high latitude areas (Sugden and John, 1976).

As faster-moving ice conduits develop, they will more efficiently transfer mass-flux from the accumulation area to the ablation zone where it is lost, especially at calving margins (e.g., lake or tidewater margins). Clapperton (1993) pointed out that because of higher temperature and precipitation, glaciers on the west side of the Andes have had a high glacial activity index, especially considering the existence of fiords and tidewater calving 'drawdown' effects (Hughes, 1998). Calving margins themselves can further facilitate ice drawdown (including from the source areas) and low surface slopes (Hughes, 1998). Fiord and valley development during the last 1 Myr on the west side of the southern Andes has been considerable (Clapperton, 1983; cf., Porter, 1989). Ice with lower surface slopes, because of dynamic fast-flowing outlet glaciers and ice streams and calving margins, may have become increasingly important over time, especially on the west side of the Andes (Clapperton, 1983).

Another aspect of landscape modification that could alter the nature of glacial dynamics is a change in bedrock elevations (Montgomery, 2002). In southern Patagonia, Singer et al. (2004b) estimated that the glacial troughs east of the mountains have deepened by >1000 m since the GPG. Great 'canyon-cutting events' appear to be a prominent feature of Quaternary Patagonian landscape development (Rabassa and Clapperton, 1990; Clapperton, 1993; Singer et al., 2004b). Lowering of elevations could lead to less positive mass balances over time (e.g., Small and Anderson, 1998; MacGregor et al., 2000). Fiords provide an example of lowering of bed elevations and overdeepenings over time (Porter, 1989). Warm-based ice eroding over broad areas may lead to areal scouring (Sugden and John, 1976), which could result in a loss of accumulation area. In alpine settings, where ice is not frozen to its bed (e.g., the middle latitude Andes), glacial erosion is an important effect in the accumulation area, e.g., as arêtes and horns develop (Tomkin and Braun, 2002). As the landscape is modified, relief may increase but the accumulation area and overall mean elevation decrease as high elevations are eroded away while low elevations expand (e.g., Small and Anderson, 1998). Alternatively, relief (i.e., 'geophysical relief') can decrease (or stay the same) if glacial erosion is concentrated near peaks and ridges (Whipple et al., 1999; Tomkin and Braun, 2002). The "glacial buzzsaw" hypothesis explains that erosion removes topography tectonically lifted above the equilibrium-line altitude (ELA), causing a relative lowering of the distributions of elevations, regardless of uplift rates and lithology (Brozović et al., 1997; Whipple et al., 1999; Tomkin and Braun, 2002). Over time, a decrease in mean elevation of the bed, in concert with the relative importance of fast-flowing ice (see above), could both work together to cause a more negative glacier mass balance, given a constant climatic forcing.

Numerous examples occur in the literature of the potential affect of changing bed elevations on glacial mass balance. For example, in many Laramide mountain ranges in the western United States, the change in mean elevation over time is estimated to be as much as 40% of the ELA change between the LGM and today (Small and Anderson, 1998). MacGregor et al. (2000) modeled (for a hypothetical system) glacial erosion causing bed lowering, ice surface lowering, and less positive mass balances. Harbor (1992) also modeled a substantial reduction (in kms) of an entire valley cross-section within 0.5 Myr, assuming no uplift (which is unrealistic for the Andes), shifting the whole simulated ice system to lower elevations.

A key tenet, if the "average" elevation of the glaciated landscape lowers, is that tectonic or isostatic compensation is not great enough to offset the effects of erosion. Various approaches to the issue (Montgomery, 1994; Gilchrist et al.1994: Tomkins and Braun, 2002; Brozović et al., 1997; Small and Anderson, 1998; Whipple et al., 1999) have concluded that during the Quaternary, the effects of glacial erosion have been relatively more important than isostatic uplift, at least in middle latitude areas. In line with the above cited studies of other areas, such adjustment (Ivins and James, 1999) in the Andes may have been limited, or more important, relatively 'outpaced' by glacial erosion. The Andes are a narrow mountain range relative to the flexural wavelength of the lithosphere (Montgomery, 1994). Glacial erosion would have occurred over a relatively narrow area, hence a reduced isostatic response may have occurred. On the other hand, the mantle/asthenosphere beneath Patagonia, influenced by subduction of segments of the oceanic Chilean spreading center, likely has an abnormal thermal regime and low viscosity (Ramos, 2005). A low viscosity has caused rapid sensitive adjustments to short term glacial changes (Ivins and James, 1999). A key question, however, is whether the subsurface characteristics are likely to induce a large persistent isostatic response over the long term relative to glacial erosion (see below, *cf.,* with areas covered by cold based ice; Stern et al., 2005).

4. Modeling

To ascertain how topographic changes to the Lago Buenos Aires (LBA) region could affect the extent of ice, we performed a simple sensitivity test (Fig. 4) using the numerical model described in Hubbard et al. (2005). The model is three-dimensional and it calculates internal deformation and basal motion based on higher-order physics (Pattyn et al., 2007). This renders it capable of dealing with high-spatial resolution, large relief and heterogeneous basal conditions. Although we use a thermomechanically coupled model, for simplicity, in these experiments ice is isothermal. Boundary conditions are required at the base of the ice sheet in the form of a 2 km digital elevation model (DEM, SRTM/ETOPO 30 data), and at the ice surface through a climate-mass balance term based on a parameterization of the ELA of current glaciers, ice caps and snow-patches distributed across the study area (Hein, 2004; Hubbard et al., 2005). The model incorporates non-local isostatic adjustment of the lithosphere to ice-loading. The mass-continuity equation is integrated forward in time with a weekly time-step and it is forced (for the purposes of this study) by shifting the parameterized ELA surface up or down, the latter to promote growth of the ice sheet. The model yields the evolving trajectory of numerous planar and diagnostic variables including thickness of the ice sheet, elevation, basal depression, stress, velocity (basal and englacial), runoff and ice-berg flux. A full description of model numerics can be found in Hubbard (2006) and its specific application along with the derivation of required data-sets relevant to this study are provided in Hein (2004) and Hubbard et al. (2005).

Three numerical experiments were performed. The first 'control' experiment was based on present topography where the ELA surface was lowered in increments until a steady-state ice sheet was simulated across the entire region that matched the GPG extent at LBA. Growth to the GPG limit required ~ 1150 m ELA depression relative to the present. Two subsequent experiments were performed where basal topography was arbitrarily 'in-filled' below the 500 m above sea level (asl) contour and the model run to a steady-state under an identical climate as the control experiment (Fig. 4). The frontal limit in these two experiments starts around the LGM position (Hubbard et al., 2005). Specifically, the topography in the two experiments was modified using a simple 'in-filling' algorithm based on fixed proportions (30% and 100%) of the residual between the actual initial elevation (E_0) and the 500 m contour, such that the new elevation (E_n) of each cell in the DEM was given by:

$$E_n = E_0 + (500 - E_0) * 30\%$$

In this manner, the lower topographic basins (i.e. the larger the above residual), the greater the modified in-filling and conversely, high mountain relief (> 500 m a.s.l.), the accumulation area and, hence, mass balance are kept constant so as to isolate the potential effect of valley development on ice extent. Though, we recognize that the 'in-fill' procedure is rather arbitrary, given the purpose of the experiment and the question being asked (what is the sensitivity of paleo-glacier extent to valley shape changes), and that no data are available to constrain realistically the true values for the Lago Buenos Aires area, our approach is pragmatic. We reiterate that no changes to the accumulation area were made and model simulations only pertain to low lying areas, i.e., below the ELA.

In-filling the low lying area of the LBA basin (and narrowing the valley) causes the modeled steady-state ice sheet to become less extensive compared to the present day topographic control (Fig. 4b). Essentially, a shallower LBA basin yields a net decrease in the basal shear stress of the outlet glacier, a corresponding reduction in ice flow and a lower overall flux (with a constant mass balance). In this particular experiment, which spans from the Pacific Ocean to the east side of the Andes, a decrease in the velocity of the outlet glacier does not yield any concomitant increase in net thickness and elevation of the LBA outlet to maintain overall balance flux (that would ultimately reduce down-profile ablation and, hence, necessitate expansion of the margin). These conditions occur because the intense 'draw-down' of the seaward-flowing outlets and ice streams draining to the west into the Pacific Ocean exert the primary control on the overall elevation of the ice sheet. Hence, the corresponding 'loss' in mass-flux from the LBA outlet glacier as a result of in-filling is readily captured by considerably steeper ice streams draining the west side of the ice divide.

The model implies that the development of Andean valleys causes more extensive ice because of more efficient through-flux and discharge. This result (i.e., bigger valleys over time = more extensive ice) is the converse of what has been observed at LBA and southern Patagonia where ice extent has decreased over time. Quite critically, however, this simulation assumes no change in accumulation area or mass balance. Hence, our conclusion based on this, albeit simple, sensitivity experiment is as follows: for ice extent to have decreased over time (e.g., the evidence in Figs. 1 and 2) the accumulation area must have also changed; the effects of glacial erosion in areas above the ELA, as outlined above, must have played an important role in producing the pattern of past glaciations.

5. Discussion

The overall extent of the ice margins has decreased throughout the last 1 Myr (Mercer, 1983; Singer et al., 2004a) in southern South America. In addition, during this time, ice coverage caused major landscape modification on both sides of the southern Andes (Clapperton, 1993; Rabassa and Clapperton, 1990; Meglioli, 1992; Singer et al., 2004b; Thomson, 2002; *cf.,* Porter, 1989). Singer et al. (2004a) hypothesized a specific link between the ice coverage trend, Andean tectonic history, and glacial erosion, which we build upon in this paper. They suggested that uplift, in part because of the subduction of several segments of the Chile Rise spreading center with the Pacific margin of South America, maximized the ice accumulation area and ice extent before or by 1.1 Ma, and subsequently glacial erosion (e.g., because of intensification of the Quaternary ice age) has reduced the accumulation area and resulted in less extensive outlet glaciers over time. In a recent summary, Ramos (2005) discussed that the climax of rapid uplift of the southernmost Andes, south of the Aysen triple junction at ~ 46°S, coincided in time and space with the collision of active oceanic Chilean spreading ridges and the South American plate (Gorring et al., 1997). Approximately west of LBA, Thomson (2002) documented relatively rapid rates of cooling and denudation between ~ 7 and 2 Ma, coeval with collision of the active mid-oceanic Chile Rise with the Peru-Chile Trench, i.e., prior to the GPG, based on fission-track thermochronology. Thomson et al. (2002) concluded that the combination of rock uplift and glacial erosion is very effective at causing localized denudation. We highlight that the pattern of GPG and subsequently less extensive ice may continue far north of the present Aysen triple junction, although it is not as well expressed (Fig. 1). Hence, although tectonic processes may vary in detail along the southern Andes, the effects of glacial erosion have broadly played a key role in development of the mountains.

High precipitation, velocities, ice flux, and calving margins could have caused faster rates of erosion on the west side of the Andes compared to the drier east side (Clapperton, 1993). This could have caused valleys and low regions to develop faster on the west side of the Andes, leading to more efficient ice drainage towards the Pacific Ocean and causing less extensive ice on the east side over the last 1 Myr. A possibility is that the east side of the Andes has become more arid over time, leading to less extensive ice. Two factors, however, counter this phenomenon as an explanation of the GPG and subsequently less extensive ice. First, although poorly dated, the pattern appears to exist on the west side of the Andes, north of ~ 42°S (Caldenius, 1932; Clapperton, 1993), where the ice margin was land based (Fig. 1). Second, and perhaps more important, the accumulation areas (and discharge) for the lobes along the eastern Andes obtain much of the snow from moist westerly precipitation (e.g., Hubbard et al., 2005; Glasser et al., 2005).

Clapperton (1990) also proposed explanations of the trend that includes tectonic subsidence, valley deepening over time caused outlet glaciers to travel less distance to discharge the same amount of ice, and climate change around Antarctica. We are not aware of any data indicating long-term tectonic subsidence, but much evidence exists for uplift (e.g., Thomson, 2002; Ramos 2005). Valley deepening is also invoked in this study; however, it is not clear whether such an effect, *on its own* without the corresponding decrease in net mass balance accompanied by lower surface elevations, would lead to decreased ice extent (Fig. 4). The equilibrium of the ice mass is dependent on the elevation of the ice surface and it may (possibly) in part adjust to deeper valleys and an increased thickness. Modifying Clapperton's (1990) idea, our conclusion is that in addition to valley deepening, high areas above the ELA also are removed.

An alternative is that climate changes explain, at least in part, the long-term trend of decreasing extent of ice. Data are scarce for southern ocean temperatures over the last 1.1 Myr near South America. Available studies, however, suggest that glacial maxima (e.g., even numbered isotope stages) around the time of the GPG were not fundamentally cooler compared to those during the last several glacial cycles (e.g., Hodell et al., 2002; Carter and Gammon, 2004). Perhaps more important, no evidence exists for a trend in increasing paleotemperatures over the last 1 Myr that would cause a decrease in the extent of ice over this time. Alternatively, precipitation may have been higher on the east side of the Andes ~1 Ma compared to the last few 100 kyr (at present, rates of precipitation on the west side are already close to the highest annual global values). No data exist for rates of paleoprecipitation to determine its relative importance compared to the topographic effects outlined in this study. On the other hand, any explanation invoking precipitation would have to explain a trend that is evident over 1500 km (~40-55˚S) and spans a range of modern climate regimes (Prohaska, 1976).

The conclusions in this paper dovetail with those in other studies. MacGregor et al. (2002) modeled over time erosion lowering the elevations of the beds and ice surfaces, leading to less positive mass balances. They highlighted a 'self-defeating mechanism' that appears in their simulations, and proposed it as an explanation for observed moraine complexes that show more extensive ice in glaciations older than the LGM. In addition, Brozović et al. (1997) and Whipple et al. (1999) emphasized that glacial erosion will limit the height of a mountain range. The former study argued specifically that irrespective of the rate of tectonic processes operating glaciations will limit the mean elevation of mountain ranges in the Himalayas, especially where precipitation is high. Here, we highlight a consequence of glacial processes limiting bedrock elevations in the Patagonian Andes. Montgomery et al. (2001) proposed that large scale climate patterns are the major control on the topographic evolution of the Andes and they highlighted the role glacial erosion has played in the southern part of the continent. Their conclusions are not necessarily in conflict with ours. The gross morphology of the Andes from the equator to the Tierra del Fuego may be influenced by broad climate and glacier variations (Montgomery et al. 2001), but non-climatic processes may still play a role in the pattern of decreasing ice extent in Patagonia.

The processes discussed here, to explain the pattern observed in Patagonia, may be relevant to other glaciated areas, especially where ice was warm-based. Smith et al. (2005) and Farber et al. (2005), for example, showed, using cosmogenic nuclide measurements, a pattern of decreasing extent of ice through the middle to late Quaternary Period in the low-latitude Andes. The trend has been documented also on other continents (e.g., Kaufman et al., 2004; Owen et al., 2005, 2006). Owen et al. (2006) described the pattern in the Himalaya and invoked decreased moisture flux into the region as the mountains have uplifted. On the other hand, they recognized that similarities may exist between the Himalayan record and an apparent global trend. We speculate that glacial erosion may play, at least in part, a role in Asian glacial extent over the long term.

A trend of decreasing glacial coverage over time is observed also in regions covered by large paleo ice sheets and cold-based ice. The Transantarctic Mountains have clear chronologic information on the timing of the maximum extent of glaciation (Marchant et al., 1993; Denton et al., 1993; Sugden and Denton, 2004). In East Antarctica, a pervasive characteristic of the glacial record is that more extensive ice occurred in the Tertiary Period, during the Miocene Epoch. Radioisotopic data on volcanic ashes interbedded and associated with the glacial diamictons and in situ cosmogenic data provide the ages. Cosmogenic dating in the Dry Valleys shows that even over the last few million years glacial extent during successive glaciations has steadily decreased, albeit in small amounts (Brook et al., 1993; Schaefer et al., 1999; Margerison et al., 2005).

In areas covered with cold-based ice, the specific combination of processes that caused decreasing ice coverage over time is expected to be different than that which caused the Patagonian trend. Isostatic rebound and uplift have been significant in the Transantarctic Mountains (e.g., Stern et al., 2005) likely because of the presence of cold-based non erosive ice. In addition, modeling work implies that greater Antarctic ice extent could be at least in part attributed to higher precipitation in a slightly warmer Miocene climate (Huybrechts, 1993).

Glacial erosion, however, has modified the Antarctic landscape over the last 11 Myr, including the pre-ice age fluvial network (Jamieson et al., 2005), and this would have affected glacial dynamics. Offshore evidence demonstrates the huge volume of material that has come from the Antarctic continent over the last 11 Myr (Anderson, 1999). Even if the accumulation area has not been lowered in Antarctica (or it has even increased), changes in the shape of the landscape and the effects described above could still be relevant. Thus, if climate changes have played a role in producing the pattern of decreasing ice extent through time in Antarctica, it would have been in concert with paleoglaciologic consequences from landscape changes.

On a final note, a general pattern of decreasing areal extent over time is not expected to be observed everywhere (e.g., West Antarctica? Anderson, 1999). In addition, Gibbons et al. (1983) argued that two or three moraines (Pinedale, Bull Lake, and in places pre-Bull Lake moraines or equivalents) are often preserved in the Rocky Mountains. This is the expected number because of a random sequence of events. Although the record in the Rockies may preserve a random sequence of events, as Gibbons et al. (1983) proposed, it is also possible that their assumption is incorrect and a trend of decreasing ice extent has occurred in the North America cordillera in the Quaternary. Recent work has highlighted that the 2-3 moraine complexes previously studied may actually represent more glaciations (Phillips et al., 1997; Kaufman et al., 2004).

6. Conclusions.

The maximum extent of glacial ice in Patagonia during the Quaternary ice age preceded global minimum temperatures and greatest ice volume, which occurred in the last ~ 800 kyr. Successive glaciations have been, in general, less extensive over the last 1 Myr (e.g., Mercer, 1983; Singer et al., 2004a; Coronato et al., 2004a). We hypothesize that the broad trend of decreasing extent of ice over time results from non-climatic effects. These include glacial erosion of mass-accumulation areas and modification of the low areas at and below the ELA. Glacial erosion over time would strongly affect the style of glaciation, especially for low-lying areas, fostering low-gradient fast ice. Furthermore, the long-term effects of glacial erosion may be evident in other areas around Earth (especially where ice has been warm-based) where a non-random trend of a decreasing extent of ice occurs with time.

Acknowledgements

We thank for discussion and insight Chalmers Clapperton, Bradley Singer, Jorge Rabassa, and Daniel Douglass. Special acknowledgment is provided for David Sugden, who influenced the paper considerably with his feedback and suggestions. Anonymous reviewers greatly clarified and improved the manuscript. Kaplan funded by the Royal Society of London and a Comer Science and Educational Fellowship. This is L-DEO contribution # 7085.

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Figures

Figure 1. (A) Map of southern South America showing schematic outlines of Quaternary glaciations, including the Greatest Patagonian Glaciation (GPG) and the last glaciation, and a schematic of the tectonic setting, including the Aysen Triple Junction (black circle) just south of 46°S. From Singer et al (2004a), Clapperton (1993), Hollin and Schilling (1981), and Caldenius (1932). (B) Digital elevation map (ETOPO 30) as a base with the limits of ice during the GPG (in black) and last glacial maximum (in red). The limits are based on the mapped extent of ice along Patagonia, south of ~38°S, from references cited in this paper. 38°S is arbitrarily chosen as the northernmost limit of digitized ice extent and the findings and conclusions of this paper are not affected by the extent of glaciers north of this latitude. The GPG covered between 558,000 and 542,000 km^2; the difference mainly results from whether the extent of ice is arbitrarily placed at the 50m (inner black dashed line) or 100m (outer black line) bathymetric contour east of Magellan and Tierra del Fuego (cf., Fig. 1a). The last glacial maximum ice covered ~ 422,000 km^2.

Figure 2. Landsat image showing the moraine record around eastern end of Lago Buenos Aires and dated lava flows (2σ) (Singer et al., 2004a). Moraine names are informal. Note the oldest glacial deposit in the field area (Telkin with arrow), the local representative of the GPG, underlies and thus is older than the Arroyo Telken Flow. The GPG was ~50 km beyond the stage 2 moraines (Kaplan et al., 2004).

Figure 3. Benthic $\delta^{18}O$ record as a proxy for global climate change. Also shown is the timing of maximum ice extent in South America (GPG). Ice volume from Shackleton et al. (1990) (i.e., mainly Northern Hemisphere ice changes).

Figure 4. Modeling simulation of valley-fill effects on the extent of ice in the LBA area. (A) Along a ~west to east profile (white line in part (B)), the ELA profile (from Hein, 2004), present bedrock topography, bedrock topography filled in by 30 and 100% (below 500 m contour, see text), and ice surfaces. The 'best fit' ELA simulates the LGM ice extent for the LBA area given no changes in topography (from Hein, 2004). (B) Map of ice sheet simulation around the latitude of LBA, including the control ice limit (yellow), and the change in ice limit after topography is filled in by 30% (pink line) and 100% (blue). On the west side the extent of ice is always controlled by the location of the continental shelf.